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THE ATMOSPHERES OF MARS, VENUS AND JUPITER

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Composition, temperature and pressure are the three parameters which determine the structure of a planetary atmosphere.

### Composition

In order to discuss the composition of a planetary atmosphere, we shall go quite far back into the history of our subject and deal first with the processes by which we believe the planetary atmospheres have evolved. By this approach one can hope to gain an understanding of the conditions which limit the structure and composition of the atmospheres of different planets.

If we assume that the planets condensed from a contracting solar nebula, then the initial composition of the planetary atmospheres would be, in general, governed by the relative abundances of elements in the sun. These are now fairly well known and the abundances compiled by Aller (1961) are plotted in Fig. 1.

The peaks in the curve reflect the stability of the respective nuclear species, as one expects for equilibrium concentrations produced by nucleosynthesis. In addition to H and He, the C, N, O group is strong; the Mg, Al, Si group is also strong, with Si the strongest; and Fe is very high. The heavier-than-iron elements fall off rapidly by 3 or more orders of magnitude, with fractional concentrations running about  $10^{-9}$  or  $10^{-10}$ .

These abundances give us an indication of the composition to be expected in the contracting cloud which formed the primitive sun. Apart from hydrogen and helium, which should be the most abundant constituents, we expect  $H_2O$ , perhaps  $NH_3$  and  $CH_4$ , oxides of C, N, Mg, Si, and Fe, all loosely joined in complex and irregular chains to make macromolecules and dust particles. It is presumably aggregates such as these which provided the original material of the planets. On this basis, it is understandable that the composition of the earth should be dominated by silicon oxides plus oxides and carbonates of Mg and Ca.

From these considerations it also seems very probable that the initial composition of the planetary atmospheres will be

dominated by

$H_2$ , He,  $CH_4$ ,  $NH_3$ ,  $H_2O$ ,  $CO_2$ ,  $N_2$ , etc.

Given this primitive composition, the atmospheres will then have developed with time through four primary processes:

- 1) The gravitational escape of gases into the interplanetary space.
- 2) The exhalation of gases from the interior of the planets.
- 3) Chemical reactions of gases in the atmosphere with the solid material of the planet.
- 4) The photodissociation and ionization of gases by the solar radiation.

Also contributing secondarily are the capture of gaseous constituents from the interplanetary medium and radioactive decay products.

Gravitational Escape of Gases. At any level in an atmosphere a number of atoms and molecules will be moving upward with

speeds in excess of that required for escape from the gravitational attraction of this planet. The critical escape velocity,  $v_c$ , from a planet is given

$$v_c = \sqrt{\frac{2MG}{R}}$$

where  $M$  is the mass of the planet,  $R$  its radius and  $G$  the universal gravitational constant. For the earth

$$v_c = 11.3 \text{ km/sec}$$

and for other planets

$$v_c = 11.3 \sqrt{\frac{M}{R}} \text{ km/sec}$$

where  $M$  and  $R$  now are in units of the earth's mass and radius. For the moon, Mars, Venus, and Jupiter the values of  $v_c$  are 2.3, 5.0, 10.4, and 61.0 km/sec, respectively.

The escape of gases can only take place effectively at high altitudes where the density and the probability of collisions are small. These requirements define the region of the atmosphere known as the exosphere, a region in which the mean free path for collisions is very large, and for a

particle moving in outward direction the probability of collision is only half. Atmospheric particles in the exosphere therefore execute ballistic trajectories in the gravitational field of the planet.

Knowing the temperature and density at the base of the exosphere, one can calculate the number of particles which will be escaping from the planet per unit time. The classical formula (e.g., Spitzer, 1952) for the time of escape,  $t_e$ , in which the density of an atmospheric constituent of molecular or atomic weight 'm' will fall to  $e^{-1}$  of its original value is given by

$$t_e = \left[ B(6\pi)^{1/2} \bar{v}/3g \right] \frac{e^y}{y} \quad (1)$$

where

$$B = \frac{n_0 T_0}{n_c T_c} \quad \text{and} \quad y = \frac{m v_c^2}{2kT}$$

$n_0$  and  $T_0$  are the particle number density and temperature respectively at the ground level,  $n_c$  is the hypothetical number density at ground if the whole atmosphere was isothermal at the escape

level temperature  $T_c$ ,  $R$  is the radius of the planet,  $\bar{v}$  is the root mean square velocity of the escaping particles, and  $v_e$  is the velocity of escape.

In the case of the earth, the average temperature of the exosphere has recently been determined to be  $\sim 1650^\circ\text{K}$ . For this value of the exospheric temperature, the time of escape of hydrogen from earth is  $\sim 1000$  years. As the age of the planets is about 5 billion years, the abundance of hydrogen in the atmosphere of the earth is comprehensible. Such gases as  $\text{CH}_4$  and  $\text{NH}_3$ , which are readily dissociated by solar ultraviolet radiation, would also lose their hydrogen, and their relative scarcity is therefore understandable.

The case of water is different. Due to the low temperature of the tropopause, almost all the terrestrial water vapor remains confined to the troposphere. The extreme ultraviolet which can dissociate water vapor does not reach the troposphere, being attenuated by ionizing and dissociative processes at the top of the atmosphere. An extremely minute amount, corresponding to the saturated vapor pressure of water at  $200^\circ\text{K}$ , will



traverse the tropopause and will be susceptible to dissociation in the upper atmosphere. This special circumstance preserves the water on our planet.

For helium the time of escape from the earth calculated from equation (1) will be of the order of  $10^7$  years. The equilibrium amount of helium in the atmosphere at any time will, however, depend both on the rate of outgassing of helium from the crust and the rate of its escape from the exosphere. Such calculations indicate that the amount of helium in the earth's atmosphere should be higher than it is observed. This discrepancy is not for the present very well understood (MacDonald, 1963).

From the above discussion it is clear that atmospheric constituents of molecular or atomic weight  $> 6$  have probably been retained by the earth up to the present time, but, as almost all hydrogen must have escaped, the present atmosphere of the earth should be in a highly oxidized state.

In Fig. 2 are plotted the times of escape of elements for Earth, Mars and Venus as a function of their atomic weights (Stewart, 1963).

These calculations have been made on the assumption that the exospheric temperature of Mars is 1100 °K (Chamberlain, 1962) and that of Venus is 2600 °K (Walker and Jastrow, 1963).

Venus, according to these considerations, must have also lost its hydrogen, while still retaining all the heavier elements. Mars loses its lighter elements at a still faster rate but the escape calculations suggest that some atomic oxygen may however be retained on Mars till the present time.

The curve for Jupiter is not shown in Fig. 2, because the time of escape, even for hydrogen, is of the order of  $10^{200}$  years. We therefore expect that the atmosphere of Jupiter and of other giant planets would still have a composition similar to that of solar atmosphere.

The atmosphere of the earth is composed of 79% nitrogen and 21% oxygen, with only traces of  $\text{CO}_2$ ,  $\text{H}_2\text{O}$  and argon.

However, in the atmosphere of another planet, which may have a similar molecular composition, the relative abundances of these molecules may be entirely different. On the earth the presence of free oxygen is probably a special circumstance which is connected with the presence of life. The  $\text{CO}_2$  abundance in the atmosphere is limited by the reaction of the gas with the crust and the formation of carbonates, while the amount of water vapor is extremely sensitive to the temperature environment of the planet.

Therefore, although from the considerations of gravitational escape of gases we may expect the same molecular composition for Mars and Venus, no predictions can be made regarding the relative abundances of these gases in these atmospheres.

The earth's atmosphere is "mixed" up to the altitude of 90 km, and therefore on an average the composition does not vary as a function of height. The mean molecular weight is approximately 29, and the total pressure at the surface is  $10^6$  dynes/cm<sup>2</sup>.

The number density at the ground is  $2 \times 10^{19}$  mol/cm<sup>3</sup>.

For an isothermal atmosphere in hydrostatic equilibrium, the variation of density with height is given by the formula

$$p = p_0 e^{-\frac{(h-h_0)mg}{kT}}$$

in which  $p$  and  $p_0$  are the densities at height  $h$  and  $h_0$ ,  $m$  is the average molecular weight per particle,  $T$  the temperature between  $h$  and  $h_0$ ,  $g$  the acceleration of gravity and  $k$  the Boltzmann constant.

The quantity  $\frac{kT}{mg}$  is the scale height,  $H$ , of the atmosphere; at an altitude of one scale height the density being reduced by a factor of  $e$ . For a mean temperature of 250 °K in the first 80 km of the atmosphere, the atmospheric scale height is  $\sim 8$  km.

Above 120 km the diffusive separation sets in and the pressure of each constituent varies in accordance with the scale height calculated for its own molecular weight.

Fig. 3 shows the change in composition in the earth's atmosphere with altitude as deduced from the recent experimental results obtained by means of satellites.

In the case of Mars, Venus and Jupiter, we expect the lower atmospheres to be "mixed" also while the upper atmospheres of these planets will be in diffusive equilibrium; the major constituent of the atmosphere at the extremity will, in each case, be hydrogen.

### Temperature

The temperature of a planet depends on its distance from the sun. In the case of the earth, the solar radiation flux, with an effective black body temperature of  $\sim 6000^{\circ}\text{K}$ , reaching the top of the earth's atmosphere, has a value of  $\sim 1.4 \times 10^6$  ergs/cm<sup>2</sup>/sec. Part of this radiation (39%) is immediately 'reflected' back to space by clouds and the atmosphere and does not play any role in the energy balance of the planet. Most of the remaining 61% of the solar radiation, mainly composed of the visible part of the spectrum, penetrates down to the

ground and heats the surface to a certain temperature denoted as the effective temperature of the planet  $(T_e)$ . In the case of fast rotating planets (e.g., Earth, Mars, Jupiter)

$$T_e^4 = \frac{S.C.}{4\sigma} (1-A)$$

where S.C. is the solar constant or the radiation flux received at the top of the atmosphere and A is the albedo of the planet over the whole solar spectrum. For the values of S.C. and albedo given above, the  $T_e$  for earth is 245 °K.

The surface radiating at this relatively low temperature emits primarily in the far infrared. A large fraction of the radiation emitted by the surface is immediately absorbed by molecules of CO<sub>2</sub> and H<sub>2</sub>O present in the lower layers of the earth's atmosphere. A part of this absorbed energy in the infrared is returned to the surface and provides extensive heating of the ground, raising the surface temperature to the observed value of ~ 290 °K.

This additional heating of the surface of ~45 °K by the return of infrared radiation from the atmosphere is referred

to as the "greenhouse effect" and is an essential element in understanding the radiation budget of the planet.

The magnitude of the "greenhouse effect" caused by a planetary atmosphere can be estimated by approximating the solution to the radiative transfer equation. The ground temperature  $T_G$  is then given by the equation

$$T_G^4 = T_e^4 (1 + 3/4 \tau) \quad (2)$$

where  $T_E$  is the effective black body temperature of the planet and  $\tau$  is the total optical thickness of the atmosphere in the infrared. This solution is based on the assumption that the atmosphere is in radiative equilibrium. In the case of earth for the observed value of surface temperature of  $290^\circ \text{K}$ , the total optical thickness of the atmosphere from equation (2) is  $\sim 2$ , which corresponds to the infrared absorption of  $\sim 86\%$ . It is fairly well known that the amount of water vapor and the  $\text{CO}_2$  in the earth's atmosphere do actually have an average opacity of  $\sim 90\%$  over the whole infrared spectrum.

In the following table are given the values of the effective black body temperatures for the other planets and also the most reliable values of the observed surface temperatures. In the case of Jupiter,  $T_G$  refers to the cloud top temperature. The differences give the magnitude of the "greenhouse effect" for each planet which in turn provide an estimate of the optical thickness of the atmosphere in the infrared.

TABLE I.

	<u>Mars</u>	<u>Earth</u>	<u>Venus</u>	<u>Jupiter</u>
$T_e$ ( $^{\circ}\text{K}$ )	209	245	235	105
$T_G$ ( $^{\circ}\text{K}$ )	230	290	600	170
G.H.E. ( $^{\circ}\text{K}$ )	21	45	370	65
Optical thickness of the atmosphere in the infrared	0.5	2.0	42	7.8

On Mars the "greenhouse effect" is very small which gives an optical thickness of the atmosphere of only 0.5, indicating that the atmosphere is optically thin; therefore, we do not expect excessively large amounts of polyatomic gases like  $\text{H}_2\text{O}$ ,  $\text{CO}_2$ , etc., which have intense absorption bands in the far



infrared. On the other hand, for Venus, the magnitude of the "greenhouse effect" is extremely large compared to the earth or Mars, and the atmosphere must be almost completely opaque to the far infrared radiation from the planet. Large amounts of polyatomic gases should be abundant on Venus.

The atmosphere of Jupiter above the clouds should have an optical thickness of  $\sim 8$  in order to explain the cloud top temperature of  $170^{\circ}\text{K}$ . We therefore also expect polyatomic gases which absorb in the far infrared at wavelengths greater than  $20\ \mu$  where most of the planetary radiation will be confined.

With these considerations in mind, we now turn to the observed properties of the atmospheres of these planets.

### MARS

The following table summarizes reliable physical data for Mars.

TABLE II.

Mean Distance from sun	1.52 A.U.
Mean Equatorial diameter	6790 km
Length of day	1.0012 earth day
Length of year	1.8808 earth year
Mass	0.1078 (earth 1)
Mean Density	3.90 gm/cm <sup>3</sup>
Gravity	377 cm/sec <sup>2</sup>
Total integrated albedo	0.26
Effective black body temperature	209 °K

#### Composition and Surface Pressure

The only gases that have been spectroscopically detected up to now in the atmosphere of Mars are carbon dioxide and water vapor. Their abundances have been estimated by several authors, but generally accepted values of the amounts of these gases in the atmosphere of Mars are not yet available. The best estimates, at the present time, vary within the following ranges:

CO<sub>2</sub> 70 ± 30 m-atm. (Kaplan, 1963)

H<sub>2</sub>O 10<sup>-3</sup> - 10<sup>-2</sup> gm/cm<sup>2</sup> (Spinrad, Munch, Kaplan, 1963;  
Dollfus 1963)

(m-atm. is the thickness of a homogeneous atmosphere in meters at normal temperature and pressure, 0 °C and 760 mm.)

Search for the presence of other gases in the atmosphere of Mars has failed to give positive results. The upper limits on their possible abundances are tabulated below.

TABLE III.

	<u>m-atm.</u>
O <sub>2</sub>	< 2.4
N <sub>2</sub> O, NO <sub>2</sub> or N <sub>2</sub> O <sub>4</sub>	< 2.0
NH <sub>3</sub>	< 0.2
CH <sub>4</sub>	< 0.1
O <sub>3</sub>	< 5 x 10 <sup>-4</sup>
SO <sub>2</sub>	< 3 x 10 <sup>-5</sup>

Earlier photometric studies by de Vaucouleurs (1954) and polarimetric measurements by Dollfus (1957) had given consistent estimates of

the surface pressure on Mars as  $85 \pm 10$  mb. Recently, however, from the analysis of pressure-dependent absorption bands of  $\text{CO}_2$  observed in the Martian atmosphere, Kaplan (1963) suggests that the total pressure at the surface of Mars may be as low as  $20 \pm 10$  mb. If this value is taken to be correct, then  $\text{CO}_2$  will make up for  $\sim 25\%$  of the total atmosphere.

The probable composition of the Martian atmosphere is therefore as follows.

TABLE IV.

<u>Gas</u>	<u>% Volume</u>
$\text{N}_2$	.72
$\text{CO}_2$	25
A	2
$\text{O}_2$	< 0.5
$\text{H}_2\text{O}$	very small

This atmosphere has a mean molecular weight of 32. With a surface pressure of 20 mb and an approximate temperature of  $210^\circ\text{K}$ , the surface density would be approximately  $5 \times 10^{17}$

particles/cm<sup>3</sup> which, for an isothermal atmosphere, should decrease exponentially with altitude with a scale height of ~ 16 km. The vertical distribution of pressure for this model is shown in Figure 4. Results of a similar computation for earth, assuming an isothermal atmosphere at 250 °K, are also shown in the figure and are in accord with recent rocket measurements. It is interesting to note that although the pressure at the surface of Mars is about 1/50 of the earth's, at an altitude of ~ 60 km the two atmospheres have the same pressure, and above this height the pressure in the Martian atmosphere is greater than the pressure at the corresponding height in the earth's atmosphere. Because the decrease of density with height of Mars is almost two times slower than in the earth, the levels of the ionosphere and thermosphere on Mars would be much higher than on the earth.

### Temperature

Planetary temperatures are usually estimated by measuring the infrared radiation emitted by the planet. Large part of the infrared spectrum is , however, absorbed by the water vapor and CO<sub>2</sub> present in the earth's atmosphere. Ground based observations

of the planets in the infrared are therefore confined to the 8-12  $\mu$  region where the earth's atmosphere is relatively transparent. This spectral region in the infrared is known as the atmospheric "window".

Extensive temperature measurements of Mars by infrared radiometry in this "window" of the terrestrial atmosphere have been made since 1926, and the results to date are quite consistent. Since the Martian atmosphere should also be largely transparent in the 8-12  $\mu$  "window", it is assumed that the radiation intensity measurements in this wavelength region refer to the surface of the planet.

Some of the earlier investigators were also able to scan the planetary disc latitudinally and longitudinally so as to obtain the diurnal and seasonal variation of temperature as a function of latitude. The up-to-date information can be summarized as follows:

	<u>°K</u>
Maximum temperature at equator	~ 300
Mean amplitude of diurnal variation	~ 60
noon to sunset (on earth in desert)	~ 30
Night side temperature cannot be measured but can probably be estimated at the equator	~ 200
Day side temperature at poles	~ 220
Mean temperature of day side	~ 260
Mean temperature of whole planet	~ 230

### Surface Features of Mars

As seen by telescope the outstanding features of Mars are (1) the dark areas (maria), (2) the general reddish-orange background ("deserts"), and (3) the polar caps.

The nature of the dark maria is not very well understood because, apart from showing fairly regular seasonal changes, they are also subject to erratic variations which make the hypothesis of earth-type vegetation somewhat doubtful. Moreover, the infrared spectrum does not have characteristic absorption bands of chlorophyll. The presence of organic material in these regions has, however, been reported by Sinton (1959) who detected C-H vibration absorptions in the  $3.5 \mu$  region which

were absent in the reflection spectrum of the desert areas.

The probability of the existence of life on Mars is controversial and beyond the scope of this review; but it will be desirable to see the observations of C-H bands on Mars repeated and probably substantiated by infrared spectroscopic measurements of the terrestrial "dark and bright" areas from an earth satellite.

The large orange areas of Mars are responsible for the reddish color of the planet seen by the naked eye. According to Kuiper (1952) they are composed of felsite rhyolite.

The polar caps are rather reliably known to be a layer of  $H_2O$  frost a few centimeters thick deposited on the surface. These caps are observed to expand to lower latitudes during fall and winter and to recede to high latitudes during spring and summer. The observed evaporation of the polar caps in summer and the almost immediate transport of the water vapor towards the winter pole across the equator is a special meteorological phenomenon peculiar to Mars. In the earth's atmosphere



the hemisphere to hemisphere mixing is extremely small.

The average speed of this "humidity wave" across the planet has been estimated to be about 45 km/day. This pole to pole circulation can perhaps only be understood by the greater length of the Martian year and the increased temperature differences between the cold and warm poles. In the summer the pole is the warmest region on the planet.

#### Blue Haze Layer

Photographs of Mars taken through blue filters indicate no surface features, and it is therefore believed to be covered with a haze layer known as "blue haze", absorbing at these wave lengths. Sometimes during the opposition, however, it clears over certain regions, and surface features become observable in the blue. These "blue clearings" last usually a few days. Conflicting arguments have been extended to explain the nature of the blue haze, but an explanation which satisfies all the observed features is yet to be found.

## Clouds

Several types of thin clouds have been frequently observed on Mars. They can be divided into three types:

A) Blue clouds, which are visible only in the blue. They are in patches and are seen near the poles and near the terminator. Polarization measurements indicate their particle size to be approximately  $0.1 \mu$ . They probably occur at altitudes less than 100 km. According to Goody (1957), they may be composed of ice crystals formed on the nuclei fed by the haze layer, while others believe them to be of the same material as the blue haze.

B) White clouds, which are visible both in the blue and yellow light. The polarization studies of these clouds suggest their nature to be the same as ice crystal clouds of a size of  $\sim 1 \mu$ . Being composed of bigger particles, they probably lie at altitudes less than that of blue clouds.

C) Yellow clouds, visible only in yellow, are very rare and variable in size. They have been seen drifting several hundred miles across the planet at a velocity of 60 km/hr. According to Goody (1957), they are composed of the same blue

absorbing material as the blue haze. Hess (1958) estimates their height to be  $\sim 6-7$  km.

### Vertical Distribution of Temperature

The atmosphere of Mars is optically thin in the infrared, and the probable absorbing gases are  $\text{CO}_2$  and  $\text{H}_2\text{O}$ , which are also present in the earth's atmosphere. The vertical temperature profile in the atmosphere of Mars, therefore, cannot be determined by observation from the surface of the earth. Only theoretical estimates exist for the vertical temperature structure of Mars. The most recent computations of this nature are due to Arking (1963), Ohring (1963) and Chamberlain (1962).

Arking has calculated a model atmosphere for Mars, allowing for convection and using the exact equation of radiative transfer for frequency independent absorption. A total optical thickness of 0.5 was chosen to obtain a surface temperature of  $235^\circ\text{K}$ , consistent with the observed mean surface temperature. Assuming an effective black body temperature of  $217^\circ\text{K}$ , an adiabatic gradient of  $-3.7^\circ\text{K/km}$  in the convection zone and an exponential

dependence of absorption on altitude with a scale height of 17 km, the temperature profile shown in Figure 4 (insert) is obtained. The convection zone is found to extend up to 8 km.

Ohring also has recently investigated the vertical temperature profile for a model Martian atmosphere containing two percent  $\text{CO}_2$ , ninety-eight percent  $\text{N}_2$  and no water vapor. The ground temperature was assumed to be 230 °K and the transfer of radiation in the atmosphere was calculated for frequency dependent absorption by the  $\text{CO}_2$ . The tropopause in this case was found to be at 9 km at a temperature of 196 °K. In the stratosphere the temperature keeps on decreasing and reaches a value of as low as 90 °K at an altitude of 42 km, where the total pressure is 2.5 mb.

Now that the mixing ratio of  $\text{CO}_2$  in the Martian atmosphere is known to be much higher, new calculations for 'non-grey' atmosphere are desirable.

The above mentioned results of the temperature distributions in the Martian atmosphere do not take into account possible

heating of the lower atmosphere by direct absorption of solar radiation in the ultraviolet. In the case of the terrestrial atmosphere, the ozone heating produces a temperature maximum at 50 km, but on Mars, with much less oxygen compared to earth, ozone would be confined to lower layers of the atmosphere. The solar radiation in the region of 2500 Å, responsible for ozone dissociation and atmospheric heating, will therefore penetrate to a much lower depth in the Martian atmosphere. Due to a higher atmospheric density at this level (compared to the density in the earth's atmosphere at 50 km) and because of the lower intensity of the solar radiation at the distance of Mars, the heating rates will probably not be as important as in the ozonosphere on earth. Still they may affect the temperature gradients substantially and limit the extent of the convection zone.

Assuming an atmospheric composition of 98%  $N_2$  and two percent  $CO_2$ , and a Martian stratospheric temperature of 134 °K, (Goody, 1957) Chamberlain (1962) has recently computed the mesospheric cooling and thermospheric heating for Mars. With the assumed model atmosphere and from the considerations of  $CO_2$  dissociation into CO and O and the consequent CO cooling at the mesopause, Chamberlain

deduces the height of the mesopause as  $\sim 130$  km at a temperature of  $76^{\circ}\text{K}$ . The considerable CO cooling at the mesopause level acts as "an effective thermostat, keeping the temperature at the exospheric or escape level (1500 km) from exceeding  $1100^{\circ}\text{K}$ ."

Fig. 5 shows the vertical distribution of temperature in the Martian atmosphere. The temperatures up to an altitude of 100 km are based on the results obtained by Arking, while for above this altitude the upper atmospheric model developed by Chamberlain has been used.

### VENUS

Venus is our nearest planetary neighbor and, after the sun and moon, the brightest object in the sky. It has therefore attracted the attention of man since the beginning of civilization. Despite the great interest, however, very little is known about the atmosphere of this planet, especially when compared with the information we have about Mars. The main reason for this deficiency is that Venus is covered with a layer of white

clouds, and the surface remains permanently invisible. Observations have, however, been made in the infrared and radiofrequency regions, and new information regarding the composition and temperature distribution in the Venus atmosphere has been obtained in the last few years. These new results of temperature and pressure at the various levels in the Venus atmosphere have forced a complete revision of ideas regarding the atmospheric structure of this planet. We shall try to summarize our present-day knowledge regarding the Venus atmosphere by first giving the physical constants of the planet and then reviewing the current hypotheses regarding its structure.

TABLE V.

	<u>Mass</u>	<u>Rad.</u>	<u>Distance</u> A.U.	<u>Density</u> gm/cm <sup>2</sup>	<u>Albedo</u>	<u>T<sub>e</sub></u> (°K)	<u>g</u> cm/sec <sup>2</sup>
Earth	1	1	1	5.5	0.39	245	980
Venus	0.81	0.97	0.72	4.8	0.73	235	842

### Composition

From the analysis of the reflected solar spectrum, the only constituent of Venus atmosphere so far established beyond question

is  $\text{CO}_2$ . Its abundance above the effective "reflecting level" of the sunlight has been estimated by several workers. Recent reinterpretations of the old spectra of Venus by Spinrad (1962) give a  $\text{CO}_2$ /atmosphere ratio of only 5% by mass.

The scattering and polarizing properties of the clouds correspond to fine droplets of  $\sim 2 \mu$  in diameter with the refractive index of water. However, Spinrad has carefully examined a high dispersion spectrogram of Venus taken by Adams and Dunham (1932) at a time when there were apparently "breaks" in the clouds, and he has concluded that the mixing ratio of water vapor down to a pressure of 8 atmospheres was less than one part in  $10^5$ .

At the same time, Dollfus (1963) has identified water vapor absorption in the upper atmosphere of Venus, and obtains a mixing ratio of  $10^{-4}$  or  $10^{-5}$  by mass, depending upon the cloud top pressure.

From the observed equality in the radar reflectivity of Venus at 68 and 12.5 cm (Muhleman, 1963), Thaddeus (1963) has recently calculated an upper limit of  $3 \text{ gm/cm}^2$  to the total amount of



water vapor which could be present in the atmosphere of Venus. From these results the atmosphere of Venus seems to be extremely dry, but a generally accepted value for the amount of water vapor present is not yet available.

Sinton (1963) has presented evidence for the existence of CO above the reflecting level of two micron photons. However, Kuiper (1963) has not detected this gas in his recent analysis of high resolution spectra of Venus.

Urey (1957) has noted that as much as 80 m-atm of oxygen could exist above the clouds without having been detected. Also according to Urey,  $\text{CH}_4$  and  $\text{NH}_3$  cannot be important constituents of the atmosphere in the presence of  $\text{CO}_2$ , and the oxides of nitrogen cannot be expected in any planetary atmosphere because they are unstable against decomposition into  $\text{N}_2$  and  $\text{O}_2$ . It seems likely that the bulk of the atmosphere is composed of  $\text{N}_2$ . Rare gases may also be present as minor constituents.

The likely composition of the atmosphere of Venus is therefore as follows:

TABLE VI.

<u>Gas</u>	<u>Percentage by Mass</u>
N <sub>2</sub>	95
CO <sub>2</sub>	5
H <sub>2</sub> O	< 10 <sup>-6</sup>

Pressure

There is conflicting evidence concerning the pressure at the cloud top level. A difference in the polarization of red and green light reflected from the cloud top has been interpreted by Dollfus (1961) as the result of molecular scattering in an atmosphere 800 m. thick at STP, which corresponds to a cloud top pressure of 90 mb. Sagan (1962) has analyzed Spinrad's results on the temperature-pressure combinations indicated by CO<sub>2</sub> lines on different plates, and he concludes that the cloud top pressure on the illuminated side lies between 0.53 and 0.83 atm. This is consistent with Kaplan's (1961) discussion of the pressure broadening of the 1.6  $\mu$  CO<sub>2</sub> bands.

From other considerations Sagan also derives a cloud top pressure of 90 mb for the night side of Venus with an uncertainty of a factor of 3.

Estimates for the pressure at the surface of Venus range from 7 atm to 200 atm. These will be discussed in more detail in a later section.

#### Temperature

Cloud top. Sinton and Strong (1960) have repeatedly measured the planetary emission from the cloud top in the 8-12  $\mu$  region and find a temperature value of  $235 \pm 10$  °K. This temperature probably refers to the cloud top level of the Venus atmosphere. Recent results from Mariner II seem to confirm this conclusion (Chase, Kaplan and Neugebauer, 1963).

From this value of temperature it is possible to construct a model for the atmosphere of Venus above the clouds. Fig. 5 shows a temperature profile computed by Rasool (1963) on the assumption that the atmosphere is in radiative equilibrium above the clouds.

The temperature should reach an asymptotic value of  $197^{\circ}\text{K}$  above  $z \sim 20$  km. The mesopause will probably occur above 60 km, and then the temperature will rise in the thermosphere owing to heating by photodissociation and photoionization reactions. Our lack of knowledge of the composition prevents us from predicting whether heating should occur below the mesopause by such mechanisms as ultraviolet absorption by ozone.

Surface. Determinations of the surface temperature of Venus by optical astronomy have been prevented by the cloud cover. However, the small amount of thermal radiation emitted by the planet in the centimeter wavelength region would penetrate through the clouds without significant attenuation, and therefore can be used to determine the temperature of the surface of the planet.

First attempts to measure radiation from Venus in the microwave region were made in 1956 by the radiotelescope of the Naval Research Laboratory. The temperature inferred from the measured radiation intensity, was, however, unexpectedly high -- of the order of  $600^{\circ}\text{K}$ , which is certainly too hot to support any imaginable form of life. Repeated measurements in the

following years have confirmed this result. (Mayer, 1961)

The observed temperatures are plotted as a function of wavelength in Fig. 7. At longer wavelengths the brightness temperature is observed to increase when the Venus-sun-earth angle becomes appreciable, indicating that the illuminated hemisphere is significantly hotter than the dark side (The 10 cm. brightness temperature may exceed  $700^{\circ}\text{K}$  on the bright side.)

Three models of Venus have been considerably discussed in connection with these measurements.

Greenhouse. First is the greenhouse model originally discussed by Sagan (1960) in which it is assumed that the  $600^{\circ}\text{K}$  radio-brightness temperature actually exists at the surface of the planet and is maintained by a very effective "greenhouse effect." The solar radiation in the visible penetrates up to the surface; the planet, thereby being heated up to a temperature  $T_e$ , emits in the infrared; but due to the presence of triatomic molecules like  $\text{CO}_2$  and  $\text{H}_2\text{O}$ , which have strong absorption bands in the infrared region, most of the radiation remains

trapped in the atmosphere and heats up the surface.

Jastrow and Rasool (1963) have recently pointed out that in such a model an extremely high infrared opacity of the atmosphere (optical thickness of 42, meaning a transmission of only  $10^{-18}$ ) will be required in order to obtain a Venus ground temperature of 600 °K.

It is very difficult to understand how the lower Venus atmosphere can be so opaque in the infrared and still be transparent in the visible. Tri-atomic molecules are good infrared absorbers, but they have "windows" at certain wavelengths which must be blocked by other molecules if the opacity is to be significant. Jastrow and Rasool have also shown that the combination of CO<sub>2</sub> and small amounts of H<sub>2</sub>O at pressure of the order of 3 atm. still contains significant windows.

However, the situation alters if the atmospheric pressure at ground level is many tens or hundreds of atmospheres. Then pressure broadening may close all the windows. Plass and Stull (1963) have shown that the optical thickness approaches 5 if the

ground pressure is about 30 atms. and the pressure broadening of  $\text{CO}_2$  alone is considered. At the high temperatures on Venus there may be many trace constituents in the atmosphere that significantly increase the opacity.

Aeolosphere. The second model, called the aeolosphere model, is due to Öpik (1961) who suggests that the blanketing must be due to dust, wind friction of the surface being the main source of energy for the high temperature. The dust is probably made of calcium and magnesium carbonates and the atmosphere is mainly composed of  $\text{CO}_2$  and  $\text{N}_2$ . No water vapor is necessarily present in the atmosphere. Due to the blanket of dust there is no sunlight penetrating to the surface. This model predicts that the microwave brightness temperatures of Venus should not change with the phase of Venus, which is apparently contradicted by the observations.

Ionospheric Emission. A third model invokes a dense ionosphere of Venus which is assumed to emit the observed microwave radiation. This model predicts that the radio observations should show a "brightening" at the limb of Venus, whereas the observations

made by Mariner II have recently indicated "limb darkening."  
(Barath et al, 1963).

### JUPITER

Relatively little is known about the structure of the atmosphere of Jupiter.

Being five times farther removed from the sun than the earth, but having a comparable albedo of 0.47, the effective temperature of the planet is only 105 °K. The high surface gravity and low temperature together suggest that the gravitational escape of gases from the planet should be extremely inefficient and the chemical composition of the atmosphere should still be primitive, containing large quantities of hydrogen and helium.

#### Composition

The only gases detected spectroscopically in the atmosphere of Jupiter are  $\text{NH}_3$ ,  $\text{CH}_4$  and hydrogen. The abundances of these



gases have been estimated by various authors, and table VII gives the approximate composition of the atmosphere of Jupiter (Spinrad and Trafton, 1963).

Helium is spectroscopically not observable from the earth and has been added due to its cosmic abundance and to account for the estimates of the cloud top pressure of 3 atms.

TABLE VII.

<u>Gas</u>	<u>Percentage by Volume</u>
H <sub>2</sub>	60
He	36
Ne	3
CH <sub>4</sub> and NH <sub>3</sub>	< 1

This table indicates that the Jovian atmosphere may not be as rich in hydrogen as expected from the discussion of gravitational escape.

Measurements of the scale height of the atmosphere of Jupiter have been made by observing the occultation of

$\sigma$ -Arietis by Jupiter. These observations yield a scale height of  $\sim 8$  km. For a mean atmospheric temperature of  $120^\circ\text{K}$ , the corresponding molecular weight would be  $\sim 4$ . This requires more helium than Table VII indicates. However, according to Gallet (1963) the relative abundance of H and He given in Table VII could also account for a mean molecular weight of 4 if the Jovian atmosphere had a substantial amount of neon.

#### Temperature.

The calculated black body temperature of Jupiter for a visual albedo of 0.47 is  $105^\circ\text{K}$ . Spectroscopic measurements in the infrared indicate a temperature of  $130^\circ\text{K}$  (Murray and Wildy, 1963) which probably refers to an intermediate level above the clouds. The cloud top temperatures have been estimated at  $\sim 170^\circ\text{K}$  (Zabriskie, 1962).

Gallet (1963) has made extensive theoretical analyses of the atmospheric structure of Jupiter below the clouds. The visible clouds have been assumed to be made of  $\text{NH}_3$  crystals, which condense at the temperature and density of the cloud top. With

allowance for the heat released in this condensation, Gallet calculates a relatively slow increase of temperature going down from the clouds to the surface and thus arrives at high densities and pressures at low temperatures. Gallet has also suggested that the observed cloud layer may be  $\text{NH}_3$  of thickness 50 km, below which there is another deck of clouds composed of  $\text{H}_2\text{O}$ . Between these two clouds is a region where  $\text{NH}_3$  rain storms can occur. In such a model the surface of Jupiter would be a few hundred km below the clouds, at a temperature of  $\sim 1000^\circ\text{K}$  and a pressure of several thousand atmospheres.

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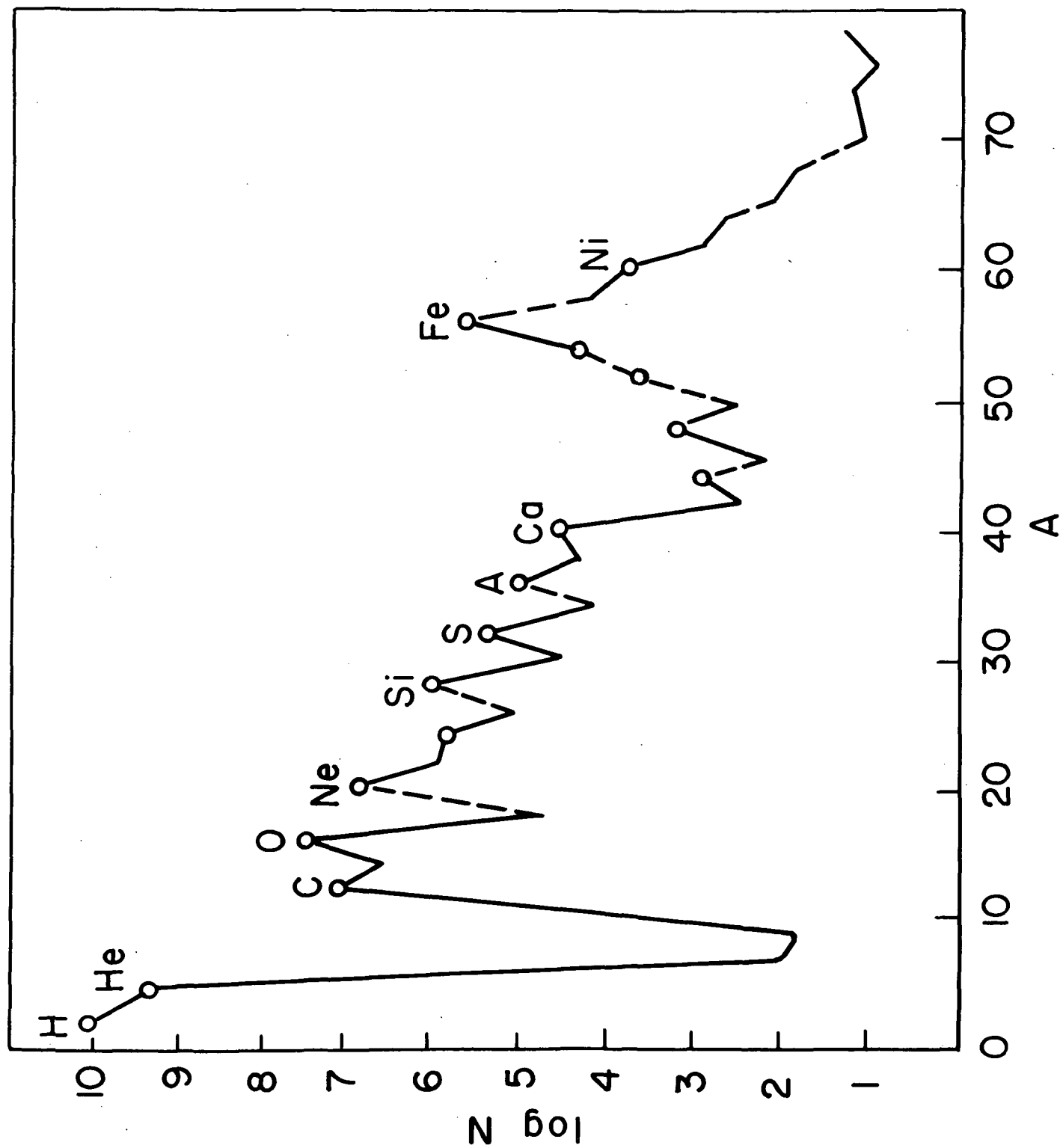
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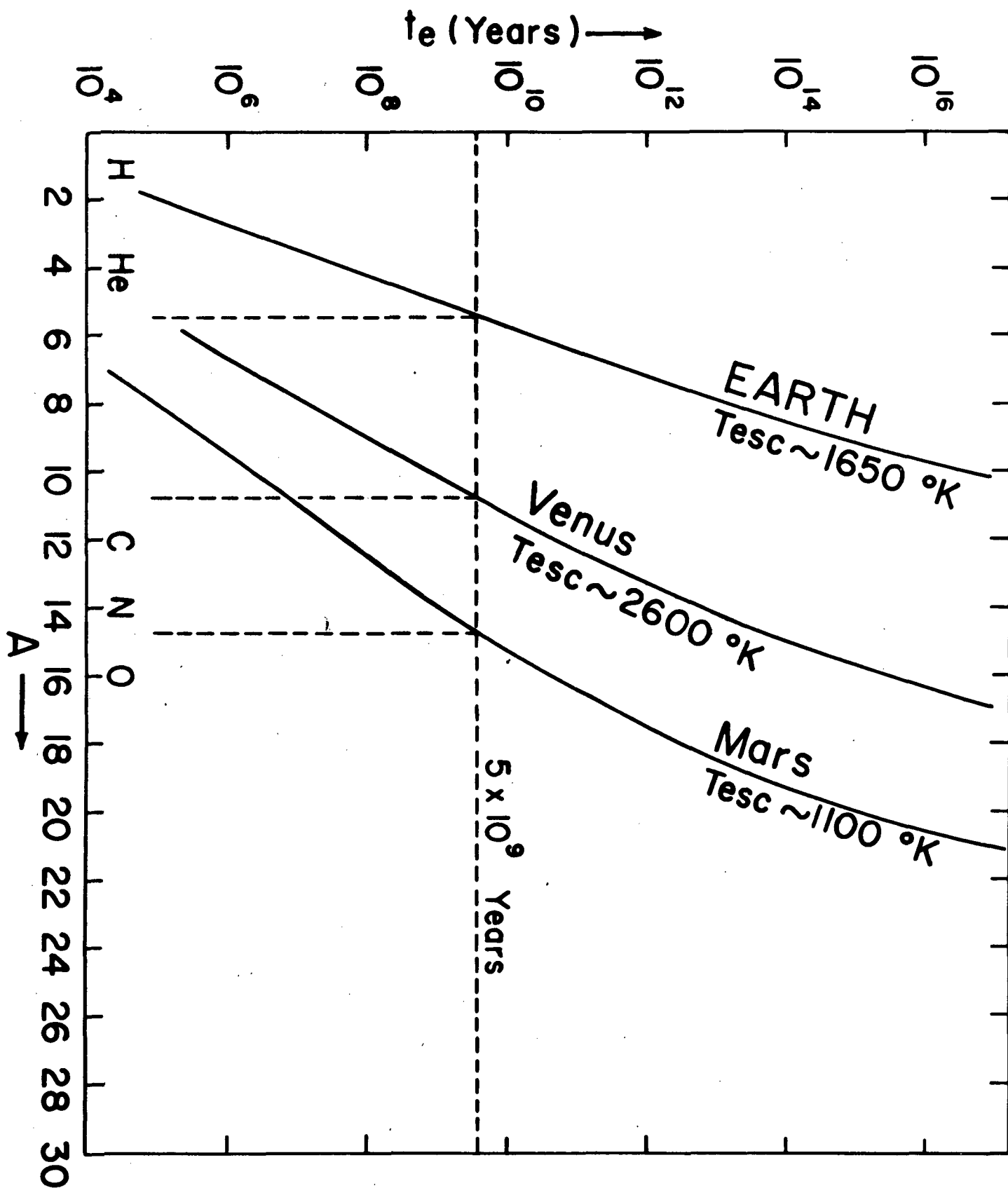
## FIGURE CAPTIONS

- Fig. 1. Abundance of the elements in the sun (after Aller, 1961): A is the atomic weight and N is the relative abundance compared to  $\text{Si} = 10^6$ .
- Fig. 2. Effective time of escape of gases of atomic weight A for Earth, Venus and Mars.
- Fig. 3. Composition of the upper atmosphere of Earth (after Jastrow, 1962).
- Fig. 4. Pressure vs. altitude for Earth and Mars. (Martian surface pressure  $\sim 20$  mb, atmospheric scale height = 16 km.)
- Fig. 5. Vertical temperature structure of the Martian atmosphere (after Arking, 1962, and Chamberlain, 1962).
- Fig. 6. Radiative temperature profile in the Venus atmosphere above the clouds (after Rasool, 1963).
- Fig. 7. Observed microwave brightness temperatures of Venus at different wavelengths.

-Fig. 1-



-Fig.2-



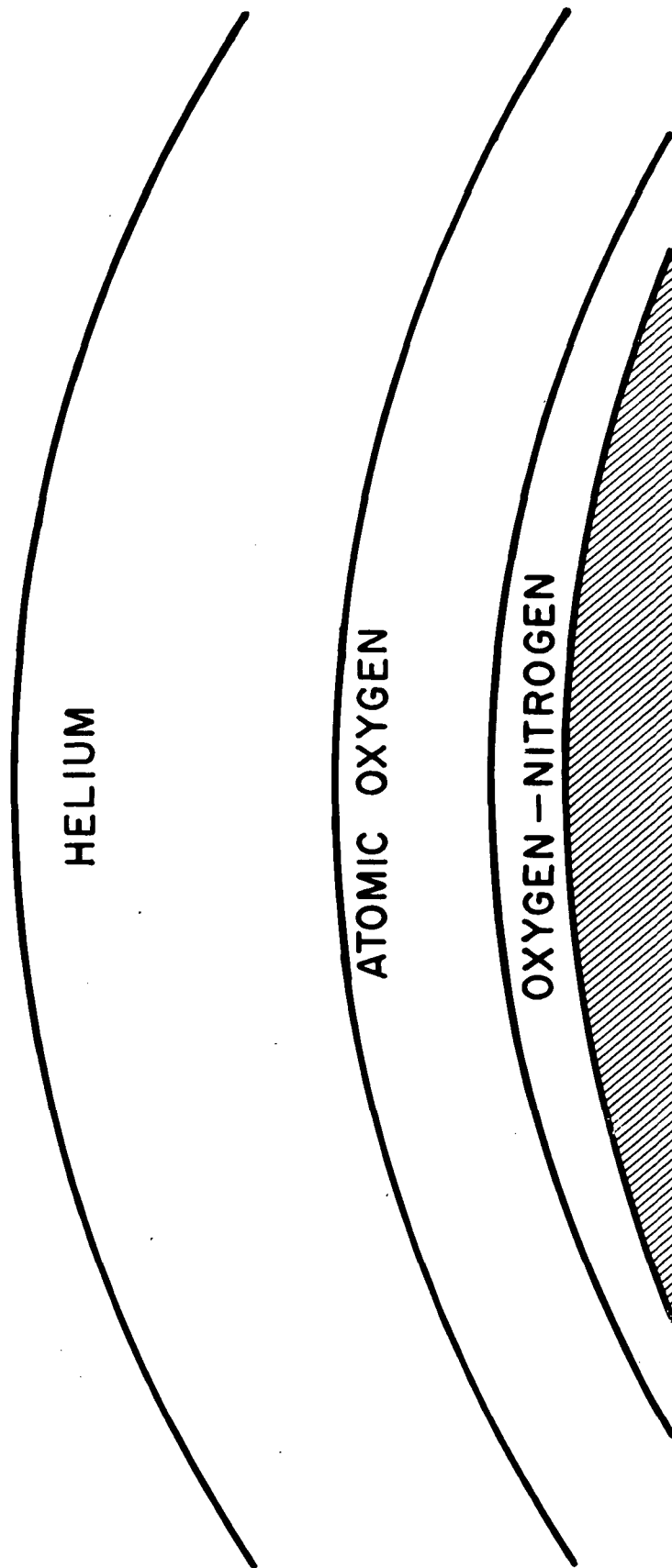
# ATMOSPHERIC COMPOSITION

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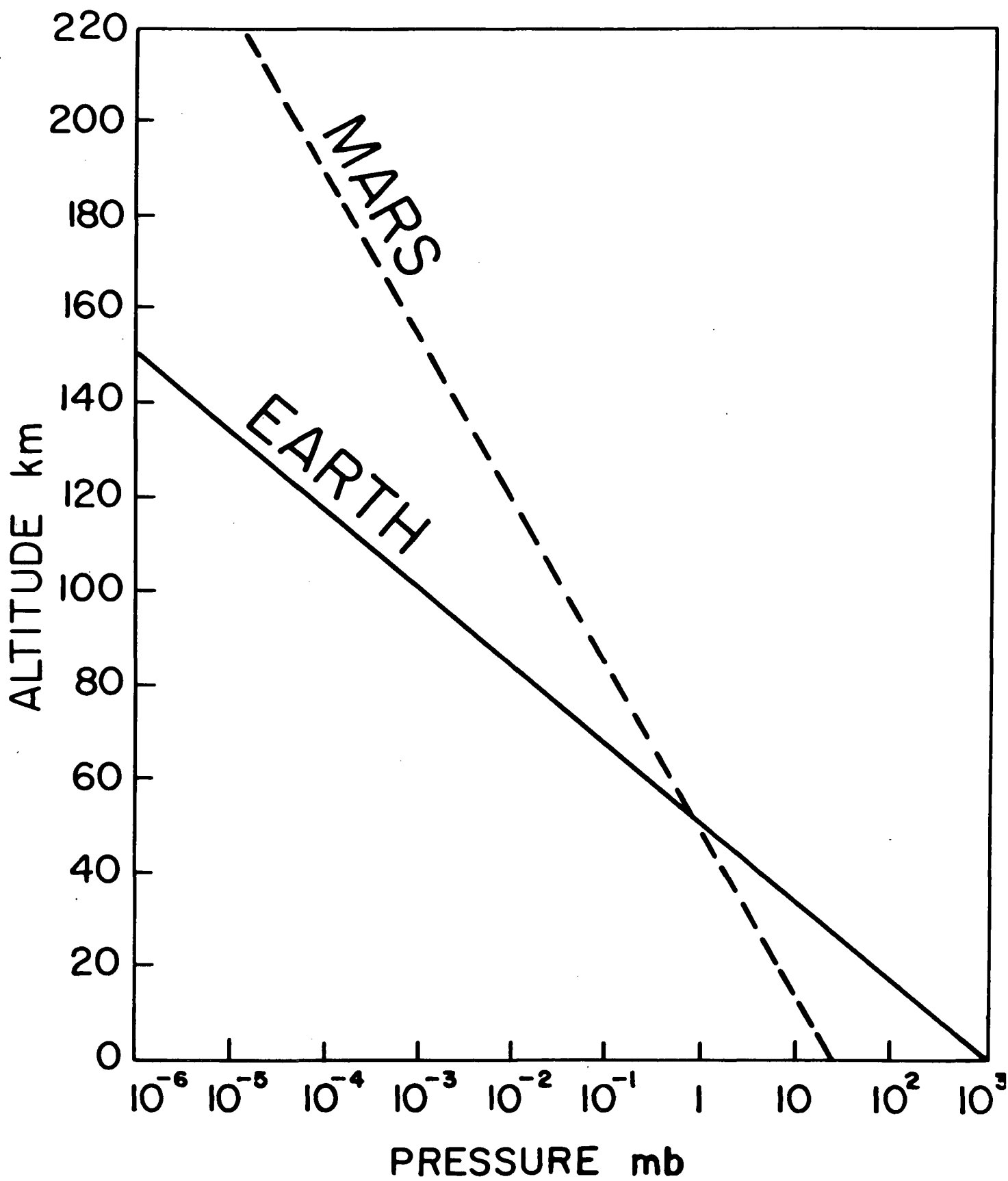
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ATOMIC OXYGEN

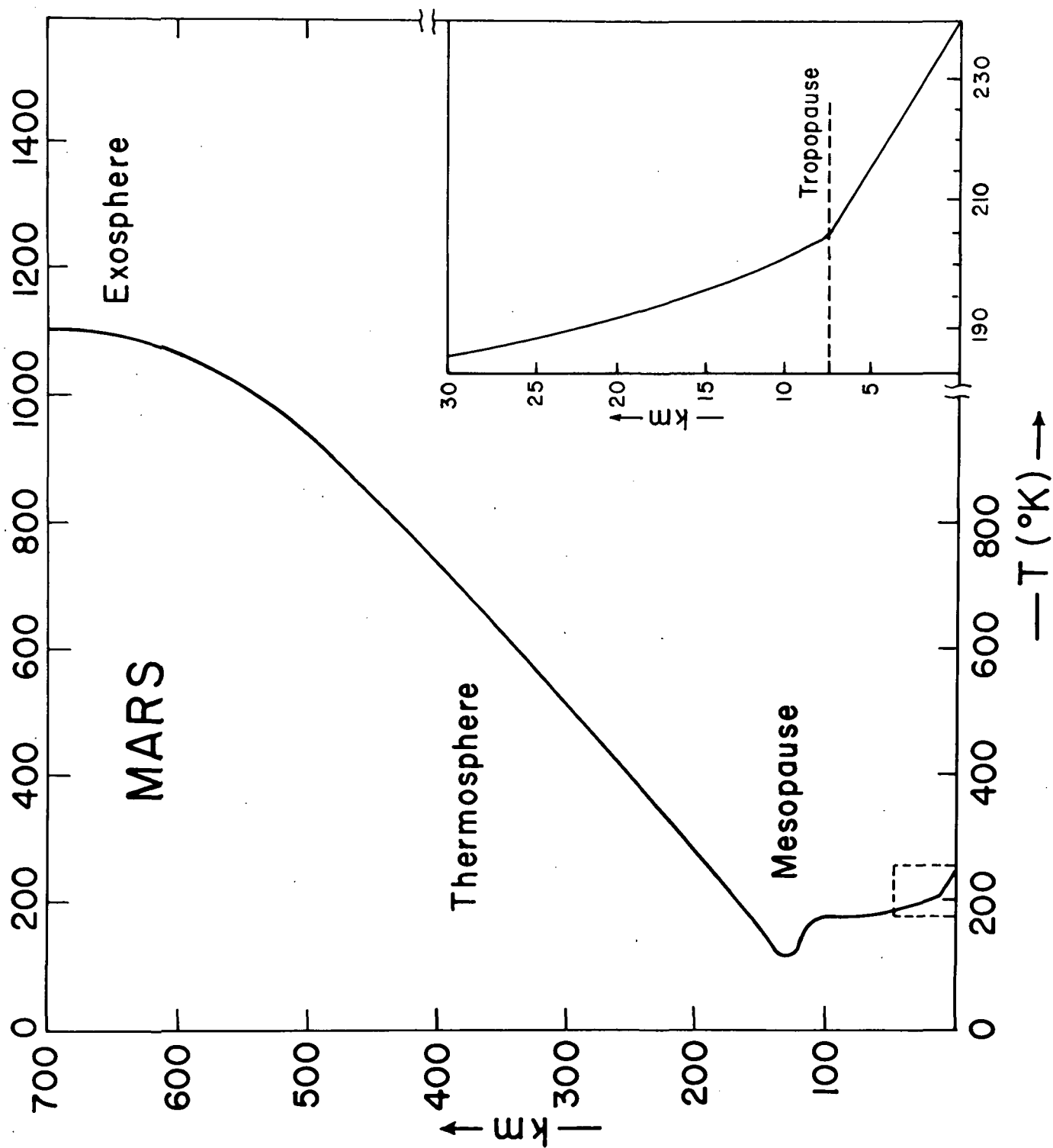
OXYGEN - NITROGEN



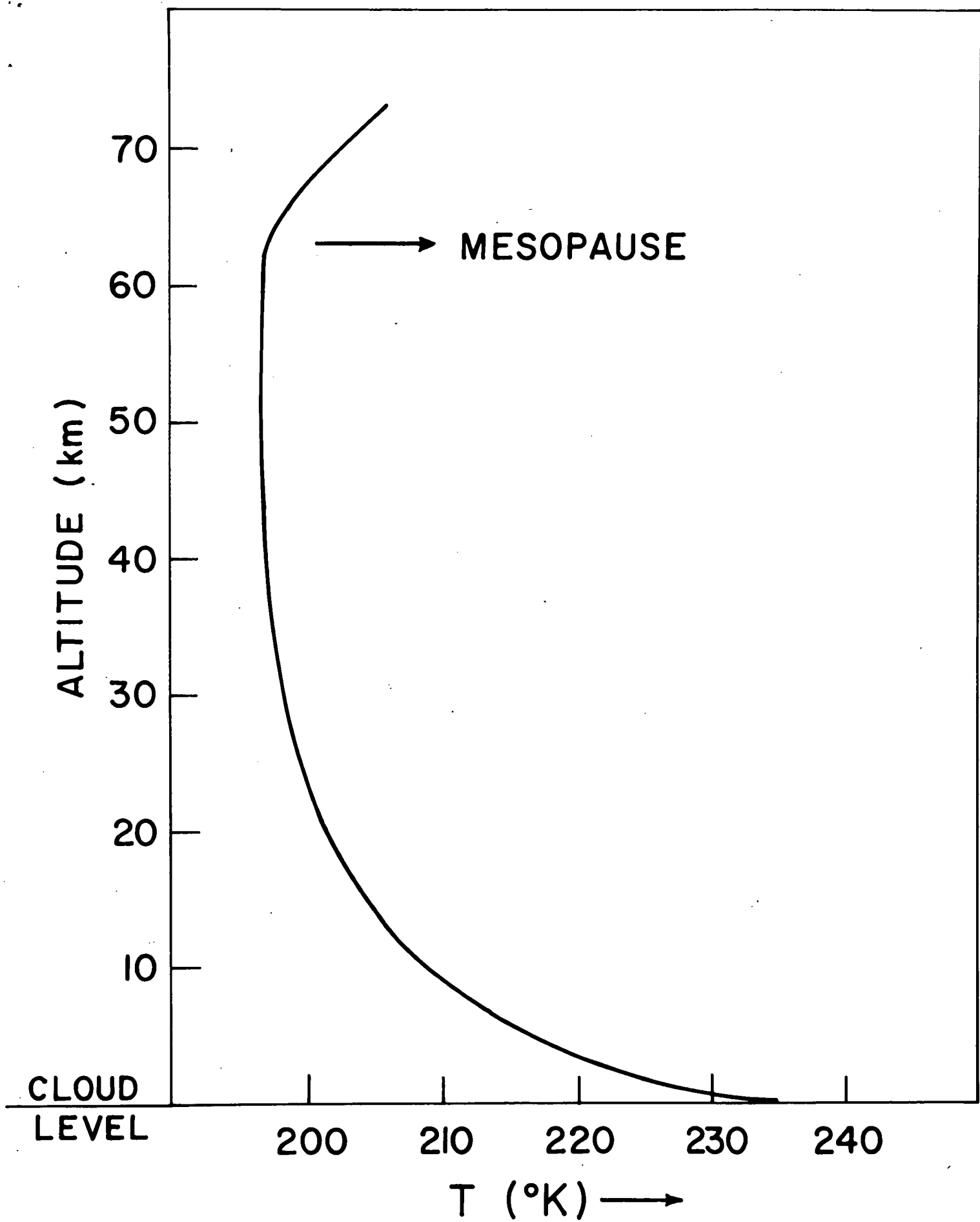
-Fig.4-



-Fig.5-



-Fig.6-



-Fig.7-

